Comment on "Piezometric response in shallow bedrock at CB1: Implications for runoff generation and landsliding" by David R. Montgomery, William E. Dietrich, and John T. Heffner

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1. Introduction

[1] Piezometric responses to rainfall on hillslopes commonly dictate the timing of landsliding. Insight to this phenomenon can be gained by evaluating the timescales for pore pressure perturbations to propagate normal and parallel to the ground surface, and these timescales can be estimated using characteristic values of hydraulic diffusivity [*Iverson*, 2000]. However, *Montgomery et al.* [2002] employed an erroneous definition of hydraulic diffusivity, leading to flawed assessment of the piezometric response timescales identified by *Iverson* [2000]. This comment aims to rectify the errors of *Montgomery et al.* [2002] and clarify the methods used by *Iverson* [2000] to estimate piezometric response timescales at the CB1 field site studied by Montgomery et al.

2. Definition and Estimation of Diffusivity

[2] For variably saturated soils (or other porous media) a suitable definition of hydraulic diffusivity, D, as stated by *Freeze and Cherry* [1979, p. 62], *Iverson* [2000], and many others, is

$$D = \frac{K}{d\theta/d\psi} \tag{1}$$

where θ is volumetric water content (volume of water per unit volume of soil or rock), ψ is pressure head, and K is hydraulic conductivity (which generally depends on θ). As defined in (1), D has the physical character of a diffusivity because $d\theta/d\psi$ represents water storage capacity [*Bear*, 1972, p. 487 ff]. The storage capacity of unsaturated soils results chiefly from filling and draining of pores, whereas the storage capacity of most saturated soils (which have grains and pore water that are effectively incompressible, and which undergo no changes in external loading) results entirely from changes in porosity (n) caused by changes in ψ, such that $d\theta/d\psi = dn/d\psi$. The quantity $dn/d\psi$ is equivalent to the specific volumetric storativity S_0 of saturated soils, as expressed by $dn/d\psi = \rho g\alpha(1 - n) = S_0$, where ρ is the pore water density (assumed constant), g is the magnitude of gravitational acceleration, and α is the soil compressibility [*Bear*, 1972, p. 202 ff.]. Thus, for fully saturated soils the hydraulic diffusivity described by (1) may be recast as

$$D = \frac{K}{dn/d\psi} = \frac{K}{S_0} \tag{2}$$

The congruity of (1) and (2) implies that hydraulic diffusivity varies continuously (and pore pressure diffusion proceeds smoothly) through the unsaturated-saturated transition in wetting soils such as those considered by *Montgomery et al.* [2002]. Typically D becomes nearly constant (and pressure diffusion becomes nearly linear) after full saturation occurs.

[3] Rather than using (1), (2), or some equivalent definition, *Montgomery et al.* [2002, paragraphs 26 and 27] defined hydraulic diffusivity as $D = KH$, where H is soil thickness. Defined in this way, D has the same dimensions (L^2T^{-1}) as D defined in (1) and (2), but it lacks the physical character of a diffusivity. Instead, the product KH is reminiscent of the aquifer transmissivity sometimes used in groundwater hydraulics. *Freeze and Cherry* [1979, pp. 58–62] elaborated the distinction between transmissivity and diffusivity.

[4] To estimate the timescales of piezometric responses in the CB1 catchment of *Montgomery et al.* [2002], a single characteristic value of D (as defined in (1) and (2)) must be identified. Although D can vary as a function of water content, pressure head, and wetting history, D typically varies less than K [*Hillel*, 1980, p. 206]. Identification of a characteristic D for CB1 requires estimation of an effective, depth-averaged storage capacity $\overline{d\theta/d\psi}$ that applies as the water table height fluctuates in response to transient rainfall. A suitable estimate of $d\theta/d\psi$ within the unsaturated zone can be obtained by inspection of soil-water retention curves for CB1 [e.g., *Torres et al.*, 1998, Figure 3]. As is typical of soil-water retention curves, the CB1 curves exhibit hysteresis and nonlinearity, but they also show that the typical magnitude of $d\theta/d\psi$ is of the order of 0.1 m^{-1} in the wet region ($\theta > 0.2$) most likely to be transiently saturated during rainfall. As soil at CB1 nears saturation (i.e., as θ exceeds about 0.3), $d\theta/d\psi$ significantly increases,

but $d\theta/d\psi$ doubtlessly decreases after full saturation is reached, owing to the change in storage mechanism summarized in (1) and (2) above. Therefore *Iverson* [2000] used the typical wet-region value $d\theta/d\psi \sim 0.1\ \mathrm{m}^{-1}$ as an estimate of the depth-averaged storage capacity $\overline{d\theta/d\psi}$ of the variably saturated soil at CB1.

[5] An independent, corroborating estimate of $\overline{d\theta/d\psi}$ can be obtained by considering specific yields of water table aquifers [*Iverson and Major*, 1987, appendix], which typically range from 0.01 for fine-grained soils to 0.3 for well-sorted, coarse granular soils [*Freeze and Cherry*, p. 61–62]. These specific yields indicate that the estimate $\overline{d\theta/d\psi} \sim 0.1\ \mathrm{m}^{-1}$ is consistent with the predominantly sandy but variable texture of the soil at CB1.

3. Estimation of Response Timescales

[6] The timescale of transient piezometric response during rain infiltration at the CB1 catchment can be estimated by combining the estimate $\overline{d\theta/d\psi} \sim 0.1\ \mathrm{m}^{-1}$ with the estimates $K \sim 1 \times 10^{-4}$ m/s and $H \sim 1$ m of *Montgomery et al.* [2002, paragraph 27]. These estimates yield $D \sim 1 \times 10^{-3}\ \mathrm{m}^2/\mathrm{s}$ and a transient piezometric response timescale $H^2/D \sim 20$ min. This matches the transient piezometric response timescale estimated by *Iverson* [2000] for CB1, and it agrees well with observations at the site. However, it differs roughly tenfold from the transient response timescale estimate of 3 hours erroneously attributed to *Iverson* [2000] by *Montgomery et al.* [2002, paragraph 26], and it differs nearly 100-fold from the 30-hour estimate obtained independently by *Montgomery et al.* [2002, paragraph 27]. It is unclear how *Montgomery et al.* [2002] obtained their 30-hour timescale estimate, as reproducing their result is problematic even if their erroneous definition of D is utilized.

[7] *Iverson* [2000] and *Montgomery et al.* [2002] also reported differing estimates of the timescale A/D that governs development of quasi-steady pore water pressures at prospective landslide sites within the CB1 catchment. The estimates differed partly as a result of differing definitions of D, but they also differed as a result of differing definitions of the planimetric contributing area A that affects quasi-steady pore water pressures at locations where landslide rupture might nucleate. As noted by *Iverson* [2000], it is impossible to identify appropriate values of A precisely, because groundwater transmits pressure upstream as well as downstream within a flow field. However, in the absence of detailed knowledge of local groundwater flow fields, it appears reasonable to use surface topography to estimate A. Therefore *Iverson* [2000] estimated A by employing the method *Montgomery and Dietrich* [1994] used to identify prospective landslide sites, whereby A equals the upslope area that hypothetically contributes surface flow to any planimetric location (x, y) where landsliding might nucleate. Of course, this methodology yields A values that vary from point to point throughout a catchment, and the timescale for quasi-steady piezometric response A/D varies throughout a catchment accordingly.

[8] To estimate a single representative A/D value for the population of prospective landslide sites within the 860 m^2 CB1 catchment, *Iverson* [2000] chose $A = 100\ \mathrm{m}^2$ but did not elaborate the rationale for this choice. The rationale was simple and was based on the observation that the largest possible A value for a landslide nucleation site at CB1 is 860 m^2 (the A value at the catchment mouth), whereas the smallest plausible A value for a nucleation site is of the order of 10 m^2 (assuming that landslides must nucleate at least a short distance downslope from the drainage divide). For the population of prospective landslide nucleation sites scattered between the drainage divide and the catchment mouth, an estimate of a typical A value is then provided by the geometric mean of 860 m^2 and 10 m^2, which is about 100 m^2. (The geometric mean provides a suitable measure of the central tendency of A values, because for discretized topography in and adjacent to CB1, the incidence of pixels with particular A values tends to decline exponentially as A values increase [*Zhang and Montgomery*, 1994, Figure 4a].) Accordingly, *Iverson* [2000] used $A = 100\ \mathrm{m}^2$ as a representative A value for CB1. A better estimate of a representative A value for CB1 could be obtained by analysis of a discretized topographic map (DEM).

[9] To evaluate A/D for CB1, *Montgomery et al.* [2002, paragraph 27] used the area of the entire catchment, $A = 860\ \mathrm{m}^2$. As a result of using this maximum possible A value and an erroneous hydraulic diffusivity, *Montgomery et al.* [2002] calculated a quasi-steady piezometric response time (A/D) of 100 days for CB1. In contrast, *Iverson* [2000] used $A = 100\ \mathrm{m}^2$ and $D = 10^{-3}\ \mathrm{m}^2/\mathrm{s}$ to estimate a quasi-steady response time of 1 day, a value more in accord with observations at the site.

4. Conclusion

[10] The chief purpose of timescale estimates like those described above is to distinguish the causes and speeds of typical piezometric responses in diverse slopes, and to thereby clarify the means by which rainfall may trigger diverse landslides [*Iverson*, 2000]. This purpose appears to have been misconstrued by *Montgomery et al.* [2002, paragraph 27], who noted that variability of soil properties at CB1 is so great that timescale estimates can vary by more than an order of magnitude, rendering precise predictions impossible. However, no method of estimation or calculation can remove the complicating effects of natural variations in soil properties, and it is not the purpose of timescale estimation to make precise predictions for a specific site such as CB1. Instead, timescale estimation serves as a starting point for more detailed quantitative analyses and as a guide for data interpretation. For example, timescale estimation leads to the inference that vertical infiltration rather than downslope groundwater flow is the principal source of pore pressure changes that trigger shallow landslides at sites like CB1 [*Iverson*, 2000]. The findings of *Montgomery et al.* [2002] corroborate this inference.

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